Met Office

The sensitivity of NAME forecasts of the transport of volcanic ash clouds to the physical characteristics assigned to the particles.

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Abstract

Volcanic ash transport and dispersion models such as NAME (Numerical Atmospheric-dispersion Modelling Environment, Met Office) require input source parameters which include the physical characteristics of the ash particles: size, shape and density. This study examines the sensitivity of NAME forecasts to the input particle characteristics used, and identifies which parameters are of first order importance to the model output. When NAME is run operationally by the London VAAC default particle characteristics are used in the absence of further information; particles are spherical with a constant density of 2300 kg m^{-3}. Considering the range of ash particle shapes and densities typically observed, the uncertainties on calculated fall velocities, when the particle shape and density are unknown, are modest, $\sim$37% and $\sim$33% respectively. These uncertainties translate into moderate changes to the travel distance of an ash particle through the atmosphere. NAME simulations of the maximum travel distance of Eyjafjallajökull 2010 ash particles show that accounting for the density distribution and measured sphericity of the ash, particles travel up to 63% further than particles with the default characteristics applied by the London VAAC. NAME simulations are found to be most sensitive to the input source particle size distribution (PSD) used, with particle shape and density having second order impacts. The input PSD used in NAME by the London VAAC should therefore be considered with care for operational simulations. Using the current London VAAC default input ash PSD, NAME simulated PSDs of the ash cloud from the eruption of Eyjafjallajökull 2010 compare well to a measured PSD of the ash cloud by the FAAM research aircraft (Turnbull et al., 2012) and a sampled deposit on the Faroe Islands (Stevenson et al., 2012), but under-estimate the mass fraction of particles with diameter $<$ 10 μm. Using measured PSDs from the eruptions of Mount Redoubt, Mount St. Helens, St. Augustine and Eyjafjallajökull the modelled PSDs also all under-estimate the mass fraction of the smallest particles when compared to the measured PSDs. The results indicate that monitoring and observations during an eruption should focus on providing total grain size distributions of the ash cloud, from which the correct source PSD to be used can be determined.

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Correction March 2, 2015

Following the publication of this Technical Report a mistake was identified in equation 14 for calculating the fall velocity of non-spherical particles. The equation, taken from Ganser (1993), should read:

$$K_2 = 10^{1.84148(-Log\Psi_R)^{0.5743}}$$

Previously the equation read:

$$K_2 = 10^{1.84148(-Log\Psi_R)0.3743}$$

The mistake has been rectified in this copy of the report, with the following changes being made.

1. The calculated uncertainty on the fall velocity of a particle when its shape is unknown has been re-calculated using the corrected sedimentation scheme and is now found to be ∼37% (previously ∼13%). Figure 2 has been updated to reflect these changes.

2. The NAME simulations of the maximum travel distance of Eyjafjallajökull 2010 ash particles which considered the sphericity of the ash were re-run using the corrected shape scheme. Ash particles with a sphericity of 0.4 are now found to travel up to 57% further than spherical particles (previously 24%). Accounting for both the sphericity and density distribution of the Eyjafjallajökull 2010 ash then particles travel up to 63% further than particles with the default characteristics applied by the London VAAC (previously 36%). Figure 6 and table 2 have been updated to reflect these changes.

3. Figure 9 which shows NAME simulations of the total column mass loadings of the Eyjafjallajökull 2010 ash cloud on the 17th May and considered the sphericity of the ash particles has also been updated. The statistical parameters calculated for these plots did not change (table 4).

4. Finally, the calculated maximum travel distance of ash particles from the eruption of Eyjafjallajökull in 2010, determined from the NAME simulated accumulated deposit over the entire eruption has been updated, and now suggests ash particles with diameter 100 μm travelled up to 1349 km from the source (figure 10). We can now account for the large particles with diameter 90 μm observed in deposits sampled in Bergen (Norway).

None of these changes have altered the conclusions of this report.

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1 Introduction

Volcanic ash represents a significant hazard to aviation; ash particles can cause serious damage to aircraft and loss of power to jet engines (Casadevall, 1994; Casadevall et al., 1996). To prevent aircraft from encountering volcanic ash clouds Volcanic Ash Advisory Centres (VAACs) are responsible for forecasting the transport and dispersion of ash in the atmosphere and issuing volcanic ash advisory messages. To do this VAACs use atmospheric dispersion models to predict the movement of ash clouds in the atmosphere, for example HYSPLIT as used by the Darwin, Wellington and Washington VAACs (Draxler and Hess, 1998; Witham et al., 2007). The London VAAC, based at the Met Office, is responsible for monitoring and forecasting the movement of ash clouds over the UK, Iceland and the north-eastern part of the North Atlantic ocean, and uses the Met Office's atmospheric dispersion model, NAME (Numerical Atmospheric-dispersion Modelling Environment) (Jones et al., 2007).

NAME is a Lagrangian dispersion model whereby a large numbers of model particles are released into, and tracked through, a model atmosphere. The modelled trajectories of the particles are due to the resolved wind field, which is provided by a numerical weather prediction model, and a stochastic element which represents turbulent mixing by unresolved eddies. The gravitational settling of heavy particles, such as volcanic ash, is also represented: particle terminal fall velocity is calculated using a Reynolds number dependent drag coefficient, particles are assumed to be spherical and the fall velocity is a function of particle diameter and particle density (Maryon, 1997; Maryon et al., 1999). Each model particle is allocated an individual diameter determined from the input particle size distribution defined as a source parameter. The mass fraction of dispersing material is divided over the model particles within each size range, and each model particle may correspond to many actual particles of a certain diameter. The exact diameter allocated to each model particle is such that the log of the diameter is uniformly distributed within each size range making up the PSD. The diameter of the model particles remains constant for the model run: aggregation is not accounted for. As well as accounting for sedimentation of particles NAME also includes parameterisations of wet and dry deposition processes. These losses are applied to each model particle, such that each can lose a proportion of its mass to the surface at every model time step. Dry deposition accounts for non-sedimentation derived deposition associated with the impaction of particles on the surface and turbulent diffusion to the surface, which is determined using a resistance analogy (Webster and Thomson, 2011). The gravitational settling of heavy particles is incorporated into dry deposition via the deposition velocity (Webster and Thomson, 2011). Wet deposition accounts for the scavenging of particles by precipitation, both within and below cloud (Webster and Thomson, 2014).

NAME does not consider any volcanic eruption near-source processes, such as buoyant erup-

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tion columns or umbrella clouds. Instead an 'effective source' is defined which aims to represent the source that is appropriate for modelling the long-range dispersion of the particles. Particles can be released over any user-defined emission profile, operationally particles are released with a uniform distribution in a vertical line source centered above the vent, ranging from the vent to the maximum plume rise height. To model the dispersion of an ash cloud NAME also requires several other input source parameters: the mass eruption rate, the plume height, the particle size distribution (PSD), particle shape and particle density are all required to initialise a model run.

There are a number of schemes available to calculate the mass eruption rate as a function of plume height and, in some cases, atmospheric conditions (e.g. Degruyter and Bonadonna, 2012; Devenish, 2013; Woodhouse et al., 2012). When NAME is used by the London VAAC the mass eruption rate (kg s^{-1}) is calculated using the empirical relationship presented by Mastin et al. (2009), which is based on observations of historical eruptions:

$$M = 141H^{4.15} \tag{1}$$

where M is the mass erupted in kilograms and H is the height of the plume above the summit in kilometers. During the eruption of Eyjafjallajökull in 2010 plume height data was provided to the London VAAC by the Icelandic Meteorological Office (IMO). Their advice was based on a range of sources including the C-band radar at Keflavik airport (Webster et al., 2012). A significant fraction of the total calculated mass (M) is expected to be transported by particles with diameter $>$ 100 μm which are not included in the operational running of NAME, as they fall out very rapidly from the plume. To represent only particles with diameter $\leq$ 100 μm it is typically assumed that 5% of the total calculated erupted mass is represented by this particle size range. There have been several studies which have explored the sensitivity of NAME simulations to the plume height and mass erupion rate defined. Dacre et al. (2013) compared NAME output mass concentrations of the Eyjafjallajökull 2010 ash cloud to measured mass concentrations by a Cloud Aerosol Spectrometer (CAS) mounted on the FAAM (Facility for Airborne Atmospheric Measurements) research aircraft. The CAS is a forward scattering (4 – 12 degrees) optical particle counter that detects particles in the nominal diameter range of 0.6 – 50 μm. Observed and NAME simulated column-integrated masses at three different locations were compared, from which Dacre et al. (2013) suggested that between 2 – 6 % of the total emitted mass from the eruption of Eyjafjallajökull in 2010 was transported by particles with diameter $<$ 30 μm. Devenish et al. (2012b) compared ground-based lidar data and NAME simulations of the ash cloud over the UK from the 16th – 18th April 2010, and suggested that 2 – 12 % of the total mass erupted survives into the distal ash cloud, greater than a few 100 km from the source. Webster et al. (2012) compared NAME predicted ash concentrations of the ash cloud from Eyjafjallajökull during April in 2010 to measured concentrations from ground lidar and a sunphotometer at Chilbolton (UK), from the Forward Scattering Spectrometer Probe (FSSP) on

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the DLR (Deutsches Zentrum für Luftund Raumfahrt) aircraft, and the CAS instrument on the FAAM research aircraft. They found that the model predictions were highly sensitive to uncertainties in plume height. Spinetti et al. (2013) consider the sensitvity of VOL-CALPUFF model forecasts of the dispersion of the plume from the eruption of Eyjafjallajökull in 2010 to the source parameterization. They also show that their model output is very sensitive to the input plume heights.

The physical charactertisitcs of ash particles, their size, shape and density, is determined by the mechanism of magma fragmentation. The complex interactions and processes driving magma fragmentation mean that particle characteristics can vary signficantly for different eruptions (e.g. Rust and Cashman, 2011). Measuring the characteristics of ash particles in an ash cloud in near-real time is challenging. Particles sampled from deposits tend to be collected close to the vent as ash is often widely dispersed, and are representative of the particle sizes which have fallen out of the near source plume only. These are dominated by the largest particles $>$ 100 μm and are un-representative of the particle size distribution remaining in the ash cloud. Measured particle size distributions from deposits are also highly sensitive to the method used to analyse the deposits (Bonadonna and Houghton, 2005). Sampling of ash clouds by instruments mounted on research aircraft are limited to areas of low ash concentration, due to safety concerns for the aircraft. Data therefore often represent the distal part of the plume, which is dominated by the smallest particles. The density of ash particles varies with particle type and particle size and measuring their density is challenging due to their small size (Bonadonna and Phillips, 2003). There are several existing models for the calculation of the terminal fall velocity of a non-spherical volcanic ash particle (e.g. Wilson and Huang, 1979; Ganser, 1993; Dellino et al., 2005), but all require the particle morphology to be characterised in the laboratory. The expressions for the terminal fall velocity are often based on particle parameters that are very difficult to measure, for example determining the particle surface area of very small (< 100 μm) vesicular particles, which is required to calculate their sphericity, is very challenging (see Bonadonna and Costa (2013) for a review). Consequently the affect of particle shape on the terminal fall velocity of ash particles is yet to be adequately described (Bonadonna and Costa, 2013). As particle characteristic data are often not available during an eruption the London VAAC has a set of default particle characteristics which are used in the absence of further information. Ash particles are assumed to be spherical and all are assigned a density of 2300 kg m^{-3}, which represents the density of rhyolitic glass (Bonadonna and Phillips, 2003). The particle size distribution used is based on observations made by Hobbs et al. (1991) of ash from explosive eruptions of Mount Redoubt, St Augustine and Mount St. Helens (table 1 and figure 1a).

We assess the sensitivity of NAME simulations of the transport and dispersion of volcanic ash clouds to the physical characteristics assigned to the ash particles. The uncertainties on the calculated sedimentation rate of the particles, given the range of particle densities and particle shapes which have been observed for volcanic ash, are determined and the impact on the modelled trans-

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Table 1: The input source particle size distribution used in NAME by the London VAAC.

Particle Diameter (μm)	Mass Fraction
0.1 – 0.3	0.001
0.3 – 1.0	0.005
1.0 – 3.0	0.05
3.0 – 10.0	0.2
10.0 – 30.0	0.7
30.0 – 100.0	0.044

port distance of the ash particles is considered. Using size-resolved output from NAME, simulated PSDs of the ash cloud and deposits from the eruption of Eyjafjallajökull in 2010 are compared to the measured PSD of the ash cloud over the southern North Sea by the FAAM research aircraft and to deposits sampled on the Faroe Islands. A sensitivity analysis of forecasts of total column mass loading of ash particles in the atmosphere to the particle characteristics assigned shows that the input PSD used has a significant impact on the forecast.

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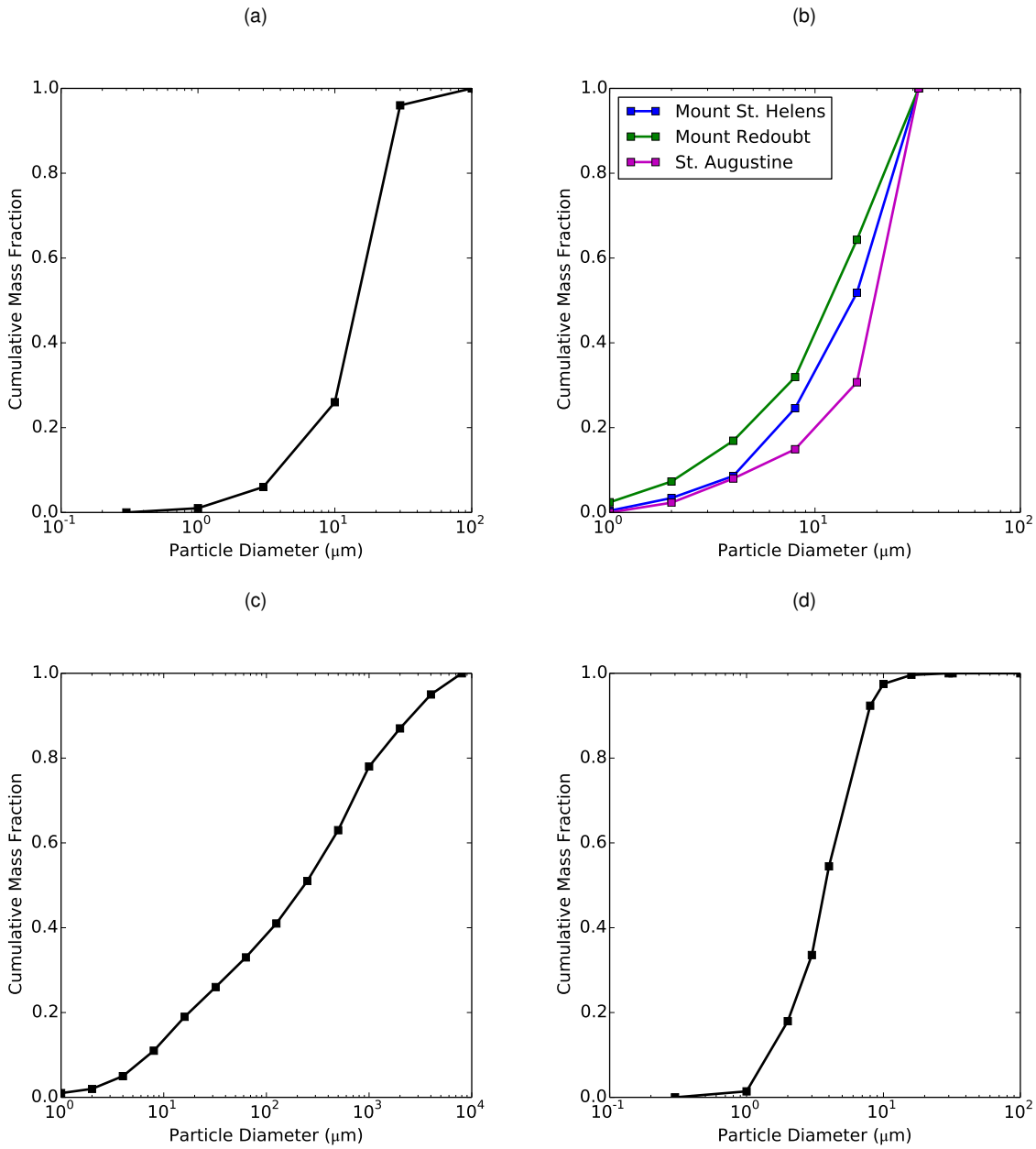

Figure 1: Input source particle size distributions (PSDs) used in NAME a. The PSD used operationally by the London VAAC. b. The measured PSDs of ash clouds from the eruptions of Mount St. Helens (Hobbs et al., 1981), St. Augustine (Hobbs et al., 1977) and Mount Redoubt (Hobbs et al., 1991) by research aircraft. c. The total grain size distribution of ash from the eruption of Eyjafjallajökull during the 4 – 8 May 2010 period based on ground measurements at distances ranging from 2 – 56 km downwind from the vent in Iceland and MSG-SEVIRI satellite measurements at distances up to 1000 km from the vent, given in Bonadonna et al. (2011). d. The measured PSD of the Eyjafjallajökull 2010 ash cloud over the southern North Sea by the FAAM research aircraft on the 17th May 2010, flight B530, taken from Turnbull et al. (2012).

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2 Sedimentation scheme in NAME

The scheme used by NAME to calculate the terminal fall velocity of ash particles in the atmosphere is outlined by Maryon (1997) and Maryon et al. (1999). The gravitational and drag forces acting on a particle falling through a fluid are given respectively by:

$$F_g = \frac{\pi D^3}{6} g(\rho_P - \rho) \tag{2}$$

$$F_a = \frac{\pi D^2}{8} C_D \rho V_{rel}^2 \tag{3}$$

where D is particle diameter, ρ is fluid density, ρ_P is particle density, C_D is the drag coefficient and V_{rel} is the magnitude of the relative velocity vector, $V_{rel} = V - V_p$ the difference between the fluid and particle velocities. Equating gravitational and drag forces (equations 2 and 3) leads to an expression for the terminal fall velocity of the particle with respect to the fluid:

$$W_T = \left(\frac{4}{3} \frac{D}{C_D} g \frac{\rho_P - \rho}{\rho} \right)^{1/2} \tag{4}$$

The drag coefficient (C_D) is, primarily, a function of the particle shape and Reynolds number:

$$Re = \frac{W_T D \rho}{\eta} \tag{5}$$

where η is the dynamic viscosity of the fluid. For spherical particles and $Re \ll 1$, Stokes flow, $C_D = 24/Re$; drag forces are linearly proportional to the relative velocity between the particle and fluid:

$$W_T = \frac{D^2}{18\eta} g(\rho_P - \rho) \tag{6}$$

For spherical particles and $Re \gg 1$ the drag force (C_D) is a non-linear function of the Reynolds number, which can be well represented by a simple empirical equation (White, 1974):

$$C_D = C_1 + \frac{24}{Re} + \frac{C_2}{1 + \sqrt{Re}} \tag{7}$$

where, for $Re < 5 \times 10^3$, $C_1 = 0.25$ and $C_2 = 6.0$. To solve for the terminal velocity of the particle (equation 4) requires an iterative approach in which the terminal velocity of a particle in Stokes flow is applied as an initial first guess.

When the size of the particle is of the same order of magnitude as the distance between gas molecules (i.e. submicron) the particle no longer moves as a continuum in the gas phase but instead as a particle amongst discrete gas molecules, which reduces the drag force acting on the particle. This is corrected for using the Cunningham Correction Factor (CCF):

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$$CCF = 1 + \alpha N_{K_N} \tag{8}$$

where

$$\alpha = 1.257 + 0.4 \exp\left(\frac{1.10}{N_{K_N}}\right) \tag{9}$$

and the Knudsen number is given by:

$$N_{K_N} = \frac{2\lambda_a}{D} \tag{10}$$

where λ_a is the mean free path of air particles:

$$\lambda_a = \lambda_{a,0}\left(\frac{P_0}{P}\right)\left(\frac{T}{T_0}\right) \tag{11}$$

Standard conditions are given by Pruppacher and Klett (1997) to be $P_0 = 1013.25$ mb, $T_0 = 293.15$ K and $\lambda_0 = 6.6 \times 10^{-6}$ cm.

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3 Sedimentation scheme for non-spherical particles

Although derived for spherical particles in the turbulent regime equation 4 is often applied to all particle morphologies and Reynolds numbers by defining an appropriate drag coefficient (Ganser, 1993). There are several schemes available in the literature which describe the drag coefficient of a non-spherical particle as a function of its shape (e.g. Wilson and Huang, 1979; Ganser, 1993; Dellino et al., 2005). The study by Alfano et al. (2011) suggests that the model of Ganser (1993) combined with sphericity (Ψ_R) as defined by Riley et al. (2003) is the best model for the calculation of the terminal fall velocity of volcanic ash, as it is highly sensitive to the effect of morphology and can be applied to all particle dimensions and Reynolds numbers typical for volcanic ash particles. The drag coefficient as defined by Ganser (1993) is given as:

$$C_D = \frac{24}{ReK_1}\left(1 + 0.1118\left[Re(K_1K_2)\right]^{0.6567}\right) + \frac{0.4305K_2}{1 + \frac{3305}{ReK_1K_2}} \tag{12}$$

$$K_1 = \frac{3}{1 + 2\Psi_R^{-0.5}} \tag{13}$$

$$K_2 = 10^{1.84148(-Log\Psi_R)^{0.5743}} \tag{14}$$

Sphericity (Ψ_R) describes the degree to which a particle approaches a spherical shape, the sphericity of a sphere is equal to 1. Sphericity can be defined in 2-dimensions as the ratio between the projected area (A) to the square of the projected perimeter (P), as described by Riley et al (2003):

$$\Psi_R = \frac{4\pi A}{P^2} \tag{15}$$

It should be noted that using this scheme with a sphericity of 1 (i.e. a sphere) does not reproduce exactly the drag coefficient used in NAME (c.f. equation 7).

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4 Uncertainties on the calculated fall velocities of volcanic ash particles

Typically the sphericity (Ψ_R) of ash particles is quoted as ranging from 0.7 – 0.9 (Folch, 2012). Bonadonna et al. (2011) measured the particle characteristics, including sphericity and density, of Eyjafjallajökull 2010 ash sampled from deposits collected on land in Iceland 2 – 56 km from the vent to the east and south-east of the volcano between the 4th – 8th May 2010. Sphericities were calculated based on a 2-dimensional analysis of images of the particles. They found Ψ_R to vary between 0.4 – 1.0, with a mean Ψ_R = 0.8, for particles in the size range 10 – 100 μm. The fall velocity of non-spherical particles with sphericities Ψ_R = 0.4, 0.7, 1.0, density 2300 kg m^{-3} and equivalent diameters ranging from 0.1 – 100 μm, falling in a constant atmosphere where atmospheric density and viscosity is taken to be that at sea level (ρ = 1.2 kg m^{-3}, $\mu = 1.82 \times 10^{-5}$ Pa s), were calculated using the drag coefficient presented by Ganser (1993) (equations 12 – 14). The maximum difference in fall velocity between a particle with Ψ_R = 1 and Ψ_R = 0.4 is ~37% (figure 2).

The density of ash varies with particle composition, texture and particle size; the smallest fragments tend to be shards of glass and are more dense than larger vesicular pumice particles (Bonadonna and Phillips, 2003; Eychenne and Le Pennec, 2012). Bonadonna and Phillips (2003) present a simple relationship to describe the density distribution of ash fragments in which the density of pumice decreases linearly with increasing size for fragments with diameters $<$ 2 mm (-1ϕ, where $\phi = -\log_2 D$) and is equal to the lithic (glass) density when the fragment diameter is $<$ 8 μm (+7ϕ). They present density distributions for rhyolitic (Askja D 1875), andesitic (Quizapu 1932) and dacitic (Hudson 1991) tephra, as shown in figure 3. The composition of the Eyjafjallajökull 2010 tephra was found to be benmoreite, trachyte and rhyolite (Gudmundsson et al., 2012). Bonadonna et al. (2011) measured the density of Eyjafjallajökull 2010 particles with diameter $>$ 8 mm by measuring their weights in air and water, and determined the density of glass, with diameter $<$ 8 μm, using a helium pycnometer. The density distribution of Eyjafjallajökull 2010 ash, using the method given by Bonadonna and Phillips (2003), is given in figure 3.

Fall velocities of spherical particles in a constant atmosphere with atmospheric viscosity and density taken to be that at sea level, and particle density distributions given for rhyolite (Askja D 1875), dacite (Hudson 1991), andesite (Quizapu 1932), Eyjafjallajökull 2010 tephra, and the default density used by the London VAAC (2300 kg m^{-3}) are calculated using the NAME sedimentation scheme (section 2). The maximum difference in the fall velocity of rhyolite, dacite and andesite tephra with the same particle diameter is ~10% (figure 4). The maximum difference in the fall velocity of rhyolite, dacite, andesite and Eyjafjallajökull 2010 tephras with the same particle diameter is ~27%. At the London VAAC particles are assumed to have a constant density of 2300 kg m^{-3}, the maximum difference in fall velocity of a given particle diameter compared to the density distributions

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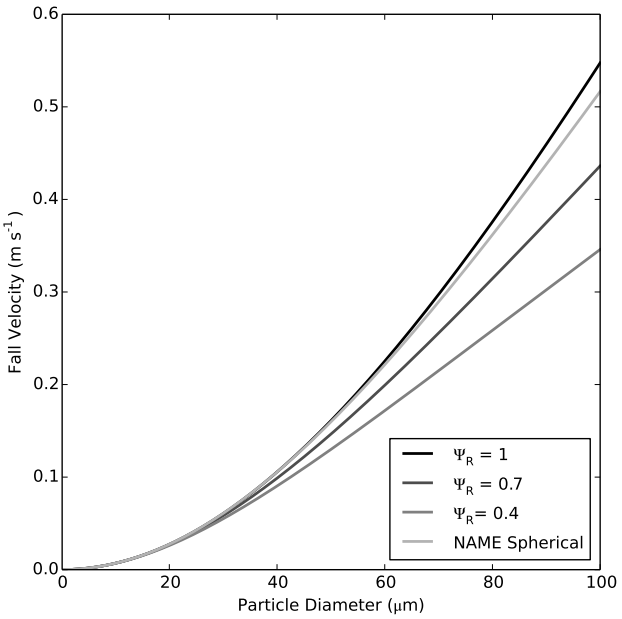

Figure 2: The fall velocity of particles with a density of 2300 kg m^{-3} and varying sphericity (Ψ_R = 0.4, 0.7, 1.0) calculated using the Ganser drag coefficient (section 3). The fall velocity of a spherical particle with density 2300 kg m^{-3} calculated using the NAME scheme for spherical particles (section 2) is compared. We assume a constant atmosphere with atmospheric density and viscosity taken to be at sea level (ρ = 1.2 kg m^{-3}, $\mu = 1.82 \times 10^{-5}$ Pa s).

given for rhyolite, dacite and andesite tephras is ∼33%. Using a constant density of 2300 kg m^{-3} over-estimates the fall velocity of the larger vesicular pumice particles. The mean difference in fall velocity of a particle with density 2300 kg m^{-3} and the Eyjafjallajökull density distribution is 6% and the maximum difference is 19%.

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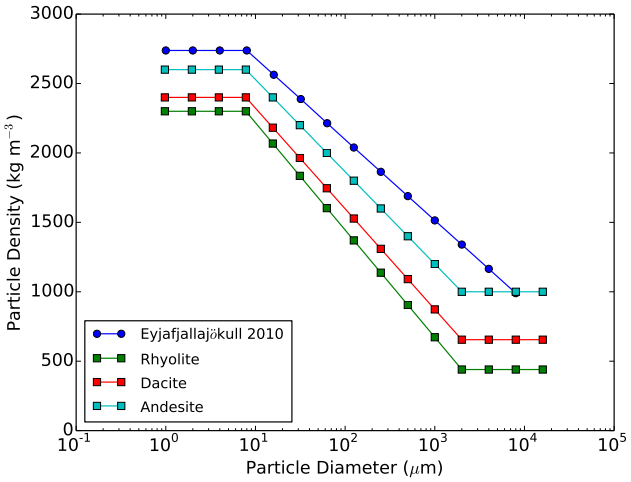

Figure 3: The density distributions of rhyoltic (Askja D 1875), andesitic (Quizapu 1932) and dacitic (Hudson 1991) tephras, as given by Bonadonna and Phillips (2003), where density is assumed to decrease linearly with increasing particle size for particles with diameter between 2 mm and 8 μm. The density distribution of Eyjafjallajökull 2010 ash, determined from the measured density of ash particles $>$ 8 mm and $<$ 8 μm collected on land in Iceland 2 – 56 km from the vent, given by Bonadonna et al. (2011), is compared. Operationally the London VAAC assumes ash particles have a density of 2300 kg m^{-3}.

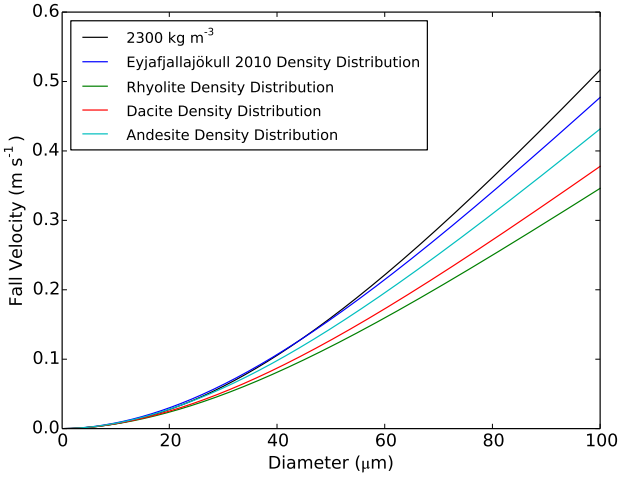

Figure 4: The fall velocity, using the NAME scheme, of spherical particles in a constant atmosphere, atmospheric density and viscosity taken to be at sea level (ρ = 1.2 kg m^{-3}, $\mu = 1.82 \times 10^{-5}$ Pa s), for varying density distributions: rhyoltic (Askja D 1875), andesitic (Quizapu 1932), dacitic (Hudson 1991) (Bonadonna and Phillips, 2003) and Eyjafjallajökull 2010 tephras (Bonadonna et al., 2011). The fall velocity of a particle with density 2300 kg m^{-3}, as used by the London VAAC, is compared.

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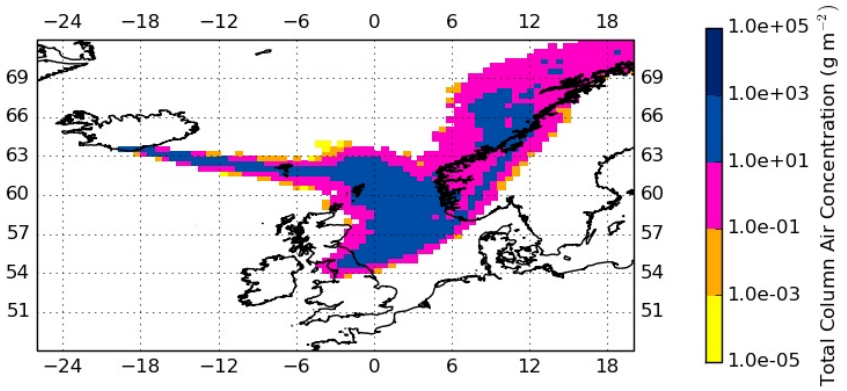

Figure 5: NAME simulated total column air concentration (g m^{-2}) of the Eyjafjallajökull 2010 ash cloud at 09:00 UTC 15/04/2010, 24 hours after the the eruption start. Particles are spherical, have a density of 2300 kg m^{-3}, and the London VAAC default input PSD was used (figure 1a). The best-guess time profile of the plume height (see figure 3 Webster et al. (2012)) is used and the mass eruption rate (M) is calculated using the Mastin relationship (Mastin et al., 2009) from which we assume 5% of the total mass is representative of the fraction of particles with diameter 0.1 – 100 μm.

5 Particle shape and density: its influence on the travel distance of volcanic ash

When considering the distance ash particles have travelled through the atmosphere for a specific event the variation in plume height with time and the weather that occured during the eruption must be accounted for, because vertical motions e.g. convective uplift and turbulent mixing, and horizontal wind velocities vary both with altitude and time. In the NAME simulations presented below the meteorological fields used are provided by the Met Office's Unified Model (UM). The global configuration of the UM is used with a temporal resolution of 3 hours and a horizontal spacing of $\sim$25 km. Plume heights are from the best-guess time profile produced using radar data (Arason et al., 2011), pilot reports (pireps) and from Icelandic coastguard observations (see figure 3 Webster et al. (2012)). Figure 5 shows the simulated total column air concentration (g m^{-2}) of the ash cloud from the eruption of Eyjafjallajökull in 2010 at 09:00 15/04/2010, 24 hours after the eruption start, during which time the plume height varied between 5.5 – 9 km asl. Particles are spherical, have a density of 2300 kg m^{-3}, and the London VAAC default input PSD is used (figure 1a).

The travel distances of the ash particles from the volcano, as a function of particle size, at 0900

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UTC 15/04/2010, are calculated as great circle distances from the output total column air concentrations (table 2). Concentrations greater than or equal to the mass of one particle, which depends on their size and density, in 1 m^2 are considered. This represents the maximum distance that one ash particle could have travelled. Travel distances of particles with the default characteristics used by the London VAAC are compared to particles with the measured characteristics of Eyjafjallajökull 2010 tephra given by Bonadonna et al. (2011) (see section 4). The maximum travel distance of a spherical ash particle with diameter 100 μm, and ρ_P = 2300 kg m^{-3}, is 456 km. When particles are assigned a sphericity of Ψ_R = 0.4 travel distances increase by up to 57%. Using the density distribution attributed to Eyjafjallajökull ash particles a 100 μm (ρ_P = 2095 kg m^{-3}) spherical particle travels $\sim$6% further, whereas the travel distance of a particle with diameter 30 μm (ρ_P = 2399 kg m^{-3}) is reduced by 1%. Accounting for both their density distribution and sphericity, the maximum travel distance of the Eyjafjallajökull ash particles is up to 63% greater than the default particles used by the London VAAC.

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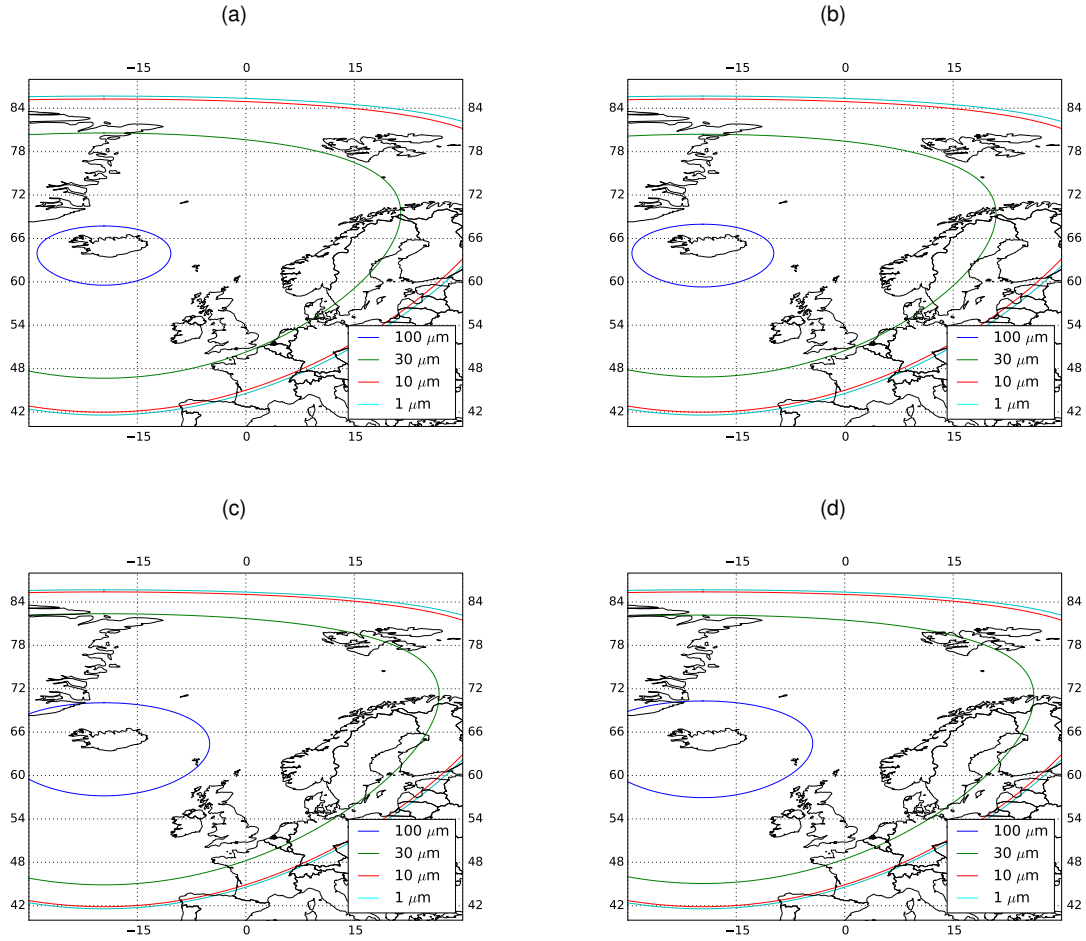

Figure 6: The maximum travel distance of ash particles from Eyjafjallajökull after the first 24 hours of the 2010 eruption. a. The particle characteristics used operationally by the London VAAC are applied: spherical particles with density 2300 kg m^{-3}. b. Spherical particles with the density distribution attributed to Eyjafjallajökull 2010 ash (figure 3). c. Non-spherical particles with Ψ_R = 0.4, the minimum measured sphericity of Eyjafjallajökull 2010 ash and density 2300 kg m^{-3}. d. Non-spherical particles with Ψ_R = 0.4 and the density distribution attributed to Eyjafjallajökull 2010 ash.

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Table 2: The maximum travel distance of NAME simulated ash particles from Eyjafjallajökull, Iceland, at 09:00 UTC 15/04/2010, 24 hours after the eruption start, using meteorology provided by the Met Office's UM. The travel distance of spherical particles with ρ_P = 2300 kg m^{-3}, as used by the London VAAC, are compared to the travel distance of ash particles with the minimum measured sphericity (Ψ_R = 0.4) and the density distribution attributed to Eyjafjallajökull 2010 ash particles. The travel distance of ash particles in a standard atmosphere with a constant wind speed (10 m s^{-1}) and plume height (10 km asl), are also shown.

	Travel Distance (km)				
	UM Met Fields Variable Plume Height				Standard Atmosphere and Constant Plume Height
Particle Diameter (μm)	London VAAC Operational	Eyjafjallajökull 2010 Density Distribution	Ψ_R = 0.4	Ψ_R = 0.4 and Eyjafjallajökull 2010 Density Distribution	Ψ_R = 0.4 and Eyjafjallajökull 2010 Density Distribution
1	2453	2453	2453	2453	864
10	2408	2408	2420	2420	864
30	1886	1866	2085	2064	864
100	456	483	716	743	248

Met Office

6 The input source particle size distribution in NAME

We now explore the sensitivity of NAME forecasts of the transport of volcanic ash clouds to the input particle size distribution used, which represents the PSD of the erupted ash at the source. The PSD of the ash cloud shifts to finer particle sizes with distance downwind due to the different sedimentation rates of the ash particles: larger particles preferentially fall out and are deposited. NAME simulates the changing PSD of the ash cloud by accounting for sedimentation. NAME simulated PSDs of the Eyjafjallajökull 2010 ash cloud and deposits are compared to a measured in-cloud PSD of the ash cloud downwind of the vent over the southern North Sea and a deposit sampled on the Faroe Islands. The effective source defined in NAME is that used by the London VAAC and described in section 1. The best-guess time profile of the plume height (see figure 3 Webster et al. (2012)) is used, and the mass eruption rate (M) is calculated using the Mastin relationship (equation 1) from which we assume 5% of the total mass is representative of the fraction of particles with diameter 0.1 – 100 μm (Webster et al., 2012). As described in Section 5 meteorological fields are provided by the UK Met Office's Unified Model (UM). Particle sizes are defined as seperate sources in the NAME input sources file, which allows size-resolved output when requesting air concentration and deposition rates.

In figure 7 the simulated PSD is compared to the measured PSD of the ash cloud sampled by the Facility for Airborne Atmospheric Measurements (FAAM) BAe-146 research aircraft during flight B530 on the 17th May 2010 1400 – 1630 UTC over the southern North Sea, flight path 52.5 – 54.5^o N, 0 – 3^o E, 3 – 8 km asl (Turnbull et al., 2012). The PSD of the cloud was measured using a Cloud and Aerosol Spectrometer (CAS) instrument which can detect particles in the size range 0.6 – 50 μm (Turnbull et al., 2012). It is important that the NAME output is at the time and location at which the ash cloud was sampled, as the PSD will depend on the meteorology, namely horizontal wind speeds and vertical wind shear, and larger particles will have preferentially fallen out of higher levels in the atmosphere. Taking the input source PSD in NAME to be that used by the London VAAC (figure 1a), results in a reasonable fit to the measured PSD of the ash cloud, (RMSE = 0.0492 table 3), but under-estimates the mass fraction of particles with diameter $<$ 10 μm and over-estimates the mass fraction of particles with diameter 10 – 30 μm (figure 7).

Hobbs et al. (1991) present measured in-cloud PSDs of ash clouds from the eruptions of Mount St. Helens, St. Augustine and Mount Redoubt using a particle size instrument mounted on research aircraft (figure 1b). The ash cloud from the eruption of Mount St. Helens was sampled on the 18th May 1980 $\sim$9 km downwind of the volcano (Hobbs et al., 1981), the ash cloud from St. Augustine was sampled on the 8th February 1976 $\sim$74 km downwind of the volcano (Hobbs et al., 1977), and the ash cloud from Mount Redoubt was sampled on the 8th January 1990 $\sim$130 km downwind of the vent. Measurements were made using a a forward light scattering particle size instrument

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which could detect particles with diameters $<$ 48 μm. Using these measured particle size distributions as the input, source, PSD in NAME, assuming a particle density of 2300 kg m^{-3}, the best fit to the measured PSD of the Eyjafjallajökull ash cloud is achieved when the St. Augustine PSD is used (RMSE = 0.0424 table 3). However, all the Hobbs et al. (1991) PSDs under-estimate the mass fraction of the smallest particles, with diameter $<$ 4 μm and over-estimate the mass fraction of ash particles in the size range 4 – 16 μm. Bonadonna et al. (2011) characterised the total grain size distribution (TGSD) of Eyjafjallajökull ash during 4 – 8 May 2010, based on both ground measurements at distances ranging from 2 – 56 km downwind from the vent in Iceland, and Meteosat Second Generation-Spinning Enhanced Visible and Infrared Imager (MSG-SEVIRI) satellite measurements up to 1000 km from the vent (figure 1c). They found that particles had a mean diameter of 250 μm and a fine ash ($D < 63$ μm) content of $\sim$ 30 wt% (figure 1c). Using this measured PSD as the input source PSD in NAME the simulated ash cloud under-estimates the mass fraction of particles with diameter $<$ 4 μm, compared to the measured PSD by the FAAM. It should be noted that if we were to use the measured PSD by the FAAM as the input source PSD in NAME then this would over-estimate the mass fraction of the smallest particles ($D < 4$ μm) at this distance downwind of the vent (figure 7).

No particles with diameter > 35 μm were detected by the CAS instrument on the FAAM aircraft during flight B530. This is replicated by all the simulated PSDs. Using the input PSD applied by the London VAAC the mass fraction of particles with diameter $>$ 30 μm is zero, and using the Hobbs et al. (1991) and Bonadonna et al. (2011) PSDs the mass fraction of particles with diameter > 16 μm is zero.

Stevenson et al. (2012) measured the PSD of accumulated Eyjafjallajökull ash deposited between the 14th April – 8th May 2010 on the Faroe Islands (62^o N, 7^o E). Particles range in diameter from $<$ 1 – 176.8 μm. Using the London VAAC operational PSD we find that the simulated PSD of the accumulated deposit on the Faroe Islands over-estimates the mass fraction of particles with diameter 10 $\leq D \leq$ 30 μm, and under-estimates the mass fraction of particles with diameter $<$ 10 μm and in the size range 30 $\leq D \leq$ 100 μm (figure 8a). Using the TGSD of Eyjafjallajökull 2010 ash given by Bonadonna et al. (2011), the mass fraction of particles $<$ 32 μm is under-estimated and the mass fraction of particles with diameter 30 – 60 μm is over-estimated (figure 8a). The measured in-cloud PSDs of ash clouds from Mount St. Helens (1980), St. Augustine (1976) and Mount Redoubt presented by Hobbs et al. (1991) and of the Eyjafjallajökull 2010 ash cloud over the southern North Sea by the FAAM aircraft (Turnbull et al., 2012) were also used as input source PSDs in NAME, however these are limited to particles with diameter $<$ 32 μm (figure 8b). Using the Mount St. Helens, St. Augustine and Mount Redoubt PSDs under-estimate the mass fraction of particles with diameter $<$ 16 μm and over-estimate the mass fraction of particles with diameter 16 – 32 μm. Using the FAAM PSD results in a poor fit to the measured PSD of the deposit, it over-estimates the

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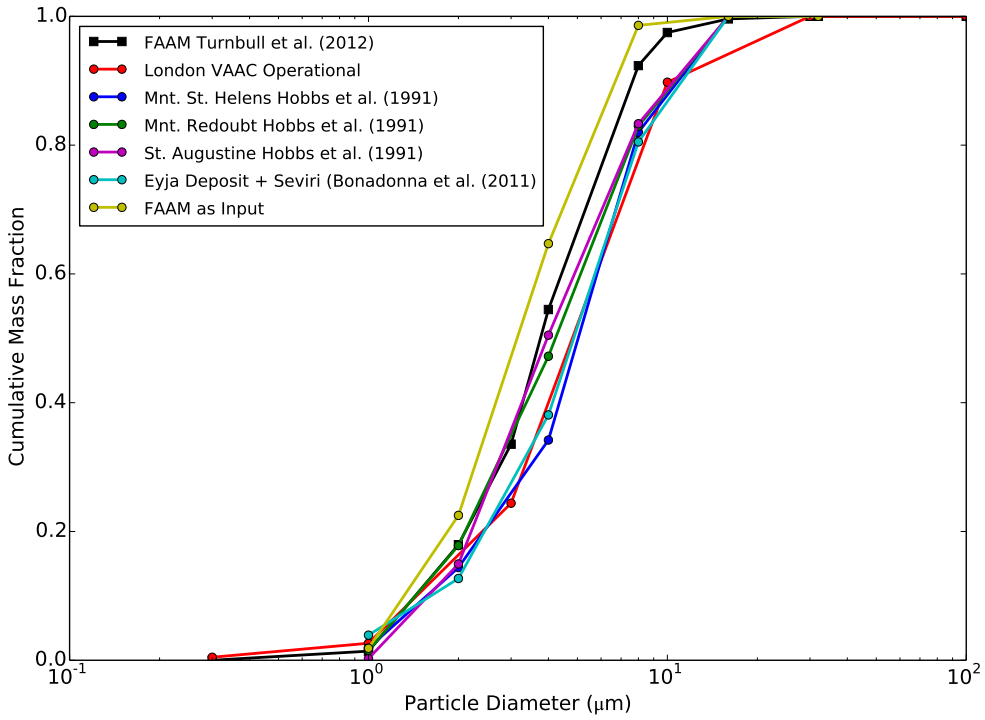

Figure 7: Comparing NAME simulated PSD of the Eyjafjallajökull 2010 ash cloud to the measured PSD by the FAAM research aircraft, flight B530 17th May 2010, 52.5 – 54.5^o N, 0 – 3^o E and 3 – 8 km asl (Turnbull et al., 2012). The input source PSDs in NAME are the London VAAC operational PSD, the measured in-cloud PSDs of ash clouds from the eruptions of Mount St. Helens 1980, Mount Redoubt (1990) and St. Augustine (1976) (Hobbs et al., 1991), the measured PSD of Eyjafjallajökull 2010 ash from deposit and MSG-SEVIRI data (Bonadonna et al., 2011). The simulated PSD of the ash cloud where the input source PSD is that measured by the FAAM at this location is also compared.

mass fraction of particles with diameter < 8 μm and under-estimates the mass fraction of particles with diameter 8 – 32 μm (figure 8b). Using the Mount Redoubt PSD results in the best fit to the measured PSD of the deposit.

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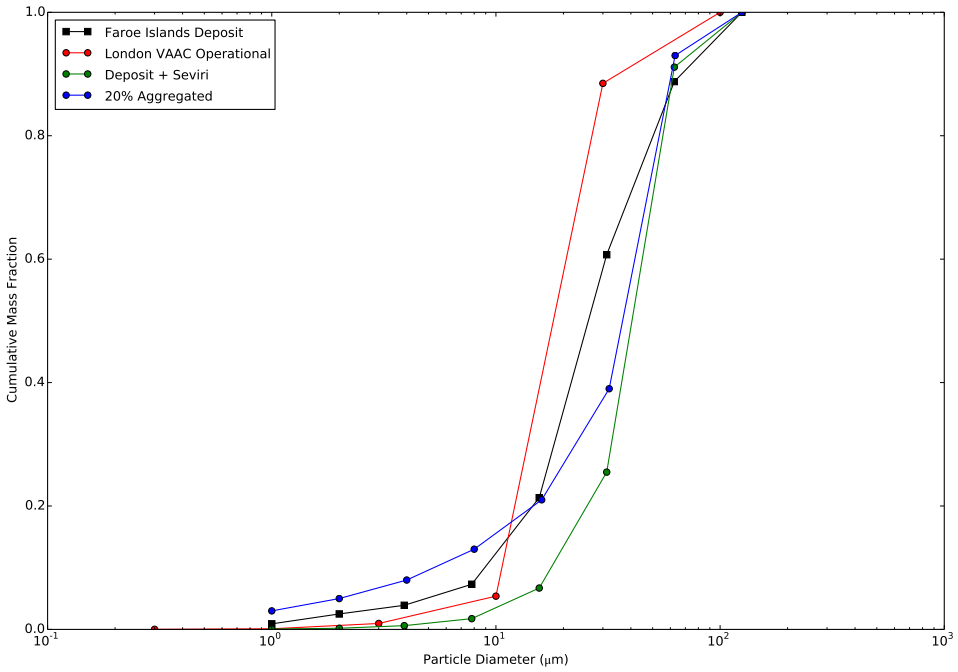

Figure 8: Comparing NAME simulated deposit PSD to the measured PSD of a deposit sampled on the Faroe Islands which accumulated between the 14th April – 8th May 2010 during the eruption of Eyjafjallajökull (Stevenson et al., 2012). a. Input source PSDs in NAME are the London VAAC operational PSD and the measured PSD of Eyjafjallajökull 2010 ash from deposit and MSG-SEVIRI data (Bonadonna et al., 2011). It should be noted that the measured deposit PSD includes particles with diameter up to 176.8 μm but here we consider particles with diameter up to 125 μm only to allow direct comparison with the source PSDs used. b. Input source PSDs in NAME are the measured in-cloud PSDs of ash clouds from the eruptions of Mount St. Helens 1980, Mount Redoubt (1990) and St. Augustine (1976) (Hobbs et al., 1991) and of the Eyjafjallajökull ash cloud over the southern North Sea by the FAAM aircraft. The PSD is limited to particles with diameter < 32 μm to allow direct comparison with the source PSDs used.

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Table 3: The root mean square error (RMSE), representing goodness of fit of the NAME simulated PSDs to the measured PSD of the Eyjafjallajökull 2010 ash cloud over the southern North Sea on the 17th May 2010 by the FAAM research aircraft, flight B530, and the measured PSD of the Faroe Island deposit, accumulated between 14th April – 8th May 2010.

Input source PSD in NAME	In-cloud FAAM PSD RMSE	Faroe Islands deposit PSD RMSE
London VAAC	0.0492	0.1056
Eyjafjallajökull Deposit and MSG-SEVIRI (Bonadonna et al., 2011)	0.0859	0.1373
Mount St. Helens In-Cloud (Hobbs et al., 1991)	0.0940	0.0706
Mount Redoubt In-Cloud (Hobbs et al., 1991)	0.0486	0.0333
St. Augustine In-Cloud (Hobbs et al., 1991)	0.0424	0.115
Eyjafjallajökull In-Cloud FAAM (Turnbull et al., 2012)	0.0523	0.401
Mount St. Helens Deposit (Carey and Sigurdsson, 1982)	0.1785	0.0751
Ruapehu Deposit (Bonadonna and Houghton, 2005)	0.0506	0.1608

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7 Implications for forecasting the transport and dispersion of volcanic ash clouds

The particle characteristics determine the fall velocity of the particle which, along with the plume (release) height and meteorological conditions, determines the maximum distance an ash particle can travel through the atmosphere before being deposited, and the maximum size of particle which can be transported to a given location. It also determines the concentration of ash particles in the ash cloud with distance downwind from the vent. In figure 9 we compare maps of modelled total column mass loading (g m^{-2}) of ash particles in the atmosphere on the 17th May 2010 at 1600 UTC, which corresponds to the time of flight B530 by the FAAM research aircraft. The particle characteristics, size, shape and source PSD, assigned in NAME are varied. A summary of statistical measures indicating the variation in the NAME simulations is given in table 4 (see Appendix A for a description of each of the statistical techniques applied).

Using the density distribution and minimum measured sphericity (Ψ_R = 0.4) of the Eyjafjallajökull 2010 ash there is very little variation from the forecast of ash dispersion across Europe produced using the London VAAC operational source parameters (figure 9 and table 4): Pearsons Correlation Coefficient (PCC) close to 1, a Fractional Bias (FB) close to zero, and a low value of the Kolmogorov-Smirnov parameter (KSP). This result is un-suprising as we showed in section 4 that the average difference in the fall velocity of spherical particles with a constant density of 2300 kg m^{-3} rather than the Eyjafjallajökull 2010 density distribution is just 6%, and there is a maximum difference of just 37% on the fall velocity of non-spherical particles with Ψ_R = 0.4 compared to spherical particles, with diameter 0.1 – 100 μm. Further, in section 5 it was shown that the maximum travel distance of volcanic ash particles with the Eyjafjallajökull 2010 density distribution in the first 24 hours of the eruption increased by a maximum of just 6% compared to ash particles with the constant density used by the VAAC, and the maximum difference in travel distance of non-spherical particles (Ψ_R = 0.4) compared to spherical particles was increased by a maximum of 57% (for the particle diameters considered: 1, 10, 30, 100 μm).

Varying the input PSD in NAME has the greatest impact on the forecast. Using the measured deposit and MSG-SEVIRI PSD, for particles with diameter up to 125 μm only (Bonadonna et al., 2011) total column mass loadings are reduced compared to using the London VAAC operational PSD, this is because the deposit and MSG-SEVIRI PSD has a higher mass fraction of larger particles which fall out of the atmosphere quicker. This results in a lower PCC, larger Fractional Bias and the lowest value of Figure of Merit in Space (FMS) because it results in a smaller plume which does not overlap completely with the operational plume. Using the PSD measured in the ash cloud on the 17th May 2010 by the FAAM aircraft the total column loading of ash in the atmosphere is higher compared to using the operational PSD. This is because the FAAM PSD has a higher mass fraction

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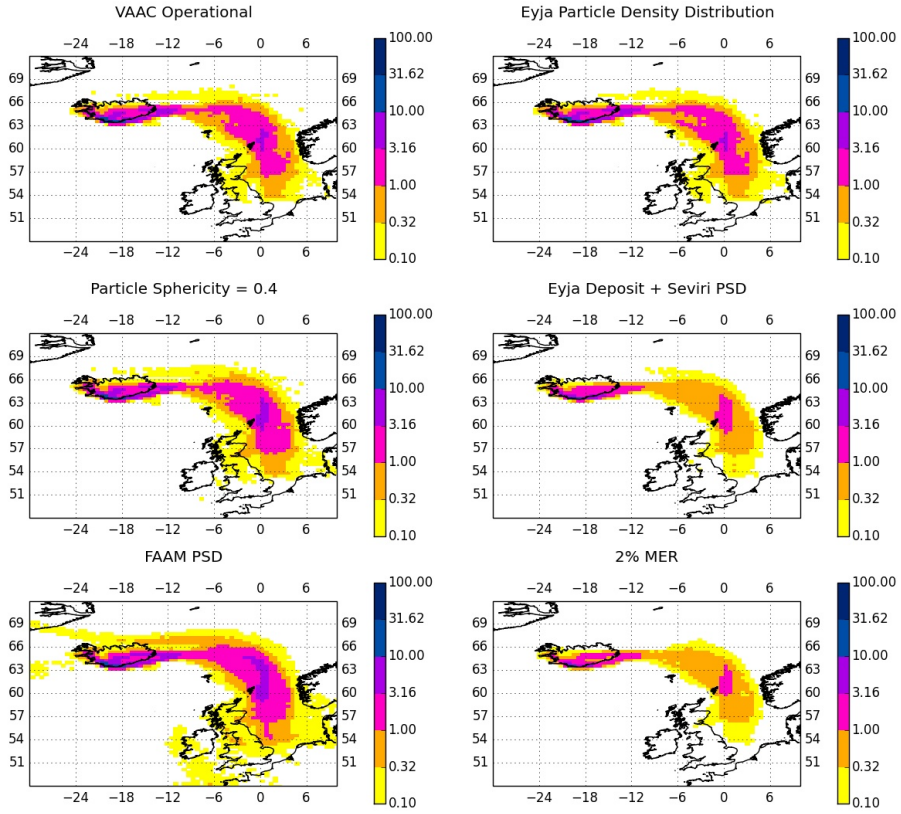

Figure 9: Comparing the NAME simulated total column mass loadings (g m^{-2}) of the Eyjafjallajökull volcanic ash cloud on the 17th May 2010 at 1600 UTC where the input operational source parameters used by the London VAAC are compared to simulations where the input particle characteristics are varied. When using the Deposit + Seviri PSD only particles with diameter up to 125 μm are considered.

of smaller particles which remain in the atmosphere for longer. Using the FAAM PSD results in good agreement for FMS because this has a higher mass fraction of smaller particles, the plume is larger and overlaps completely the plume forecast using the operational PSD.

During the Eyjafjallajökull 2010 eruption the London VAAC assumed that 5% of the total calculated mass, using the Mastin et al. (2009) relationship (equation 1), represented particles with diameter 0.1 – 100 μm. Taking 2% of the total calculated mass, within the range of 2 – 6% suggested by Dacre et al. (2013) to be representative of small ($<$ 30 μm) particles in the distal Eyjafjallajökull plume, results in a similar impact to varying the input source PSD. There is a perfect positive correlation (PCC =1) with the operational output and good aggreement in FMS values, as there is no spatial difference in the plumes only a change in magnitude. There is a large Fractional Bias when compared to the operational output due to the lower total column mass loadings.

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Table 4: The Pearson Correlation, Fractional Bias, Figure of Merit in Space and Kolmogorov-Smirnov Parameter for NAME simulations of total column mass loading, where the input particle characteristics are varied, see Appendix A for definitions of the statistical techniques used.

	Pearson's Correlation Coefficient	Fractional Bias	Figure of Merit in Space	Kolmogorv-Smirnov Parameter
Operational Vs.				
Sphericity Ψ_R = 0.4	0.997	0.11	98.42	0
Eyjafjallajökull 2010	0.999	-0.029	98.37	1
Density Distribution Deposit and MSG-SEVIRI PSD	0.971	-0.56	98.19	3
FAAM PSD	0.975	0.517	98.45	2
2% Mastin MER	1.0	-0.86	100	2

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8 Discussion

Given the range of observed particle shapes and densities the uncertainties on calculated fall velocities of volcanic ash when the particle characteristics are not known are modest. Considering a constant density of 2300 kg m^{-3} and andesite, rhyolite and dacite density distributions the terminal fall velocity of a particle with a diameter between 0.1 – 100 μm varies by up to 33%. The difference in terminal fall velocity of a spherical particle and a particle with sphericity Ψ_R = 0.4, with diameter 0.1 – 100 μm, is up to 37%. These uncertainties translate to moderate increases in the calculated travel distance of volcanic ash; accounting for the end-member sphercity (Ψ_R = 0.4) and density distribution of Eyjafjallajökull 2010 ash a particle with diameter 100 μm travelled up to 743 km, a spherical ash particle with the VAAC default density (2300 kg m^{-3}) travelled up to 456 km (table 2). This suggests that using a nominal particle density in NAME when no data on the particle characteristics are available is reasonable during operational practice. Our findings agree with the work of Scollo et al. (2008) who conducted a sensitivity analysis of the two dimensional advection-diffusion model TEPHRA. They simulated two sub-plinian eruptions: the 22nd July 1998 eruption of Etna (Italy) and the 17th June 1996 eruption of Ruapehu (New Zealand). Input factors considered in the sensitivity analysis included the density of lithics and pumice assigned as input parameters. The lithic and pumice densities were varied between 2300 – 2500 kg m^{-3} and 500 – 2300 kg m^{-3} respectively. Model output of accumulated mass per unit area was compared to tephra samples collected from the Etna and Ruapehu deposits. For the two test cases they found that the particle density can be fixed at a nominal value as varying the density has a negligable affect on the model predicitions

Stevenson et al. (2015) compare two different schemes to calculate particle terminal fall velocity to a measured fall velocity of a rhyolitic ash sample (Ash Hollow) given by Riley et al. (2003). The first uses Stokes settling law, assuming spherical particles with a density of 2300 kg m^{-3}, the second uses the Reynolds number drag coefficient given by Ganser (1993) which accounts for the non-spherical shape of the particles. Using the NAME + Ganser scheme they calculate the maximum travel distance of rhyoltic ash particles using the measured particle sphericity of the Ash Hollow sample (Ψ_R = 0.7) (Riley et al., 2003) and the density distribuion given for rhyolitic ash (Askja 1875) by Bonadonna and Phillips (2003). Particles are released at a height of 10 km above sea level (asl), the maximum height of the plume during the eruption of Eyjafjallajökull in 2010 (Gudmundsson et al., 2012), and a constant horizontal wind speed of 10 m s^{-1} is assumed; the mean wind speed over Iceland at 500 hPa ($\sim$ 5.5 km asl) during the 2010 eruption of Eyjafjallajökull was 12 m s^{-1} (Petersen et al., 2012). They show that ash particles with a diameter of 50 μm remain aloft for 24 hours and particles with a diameter of 33 μm can reach London from Iceland. Using the measured terminal fall velocity of rhyolitic (Ash Hollow) particles resulted in a significant increase in the predicted travel distance, particles with an equivalent area diameter of 50 μm travelled up to

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3$\times$ further, although they note that uncertainties on the measured length scales of the Ash Hollow particles are not known. Stevenson et al. (2015) show that the calculated travel distances of the particles are compatible with observations of grainsize distributions from deposits of Icelandic cryptotephra sampled accross NW Europe. They conclude that ash particles with diameters $>$ 12 μm, which can not be correctly identified by Brightness Temperature Difference methods used to identify ash clouds from satellite data, can travel significant distances from the volcano; considering the measured fall velocity of rhyolitic ash a particle with diameter 80 μm can travel 850 km in 24 hours. Using the approach of Stevenson et al. (2015), assuming a constant release height (10 km asl) and wind speed (10 m s^{-1}), and considering Eyjafjallajökull 2010 ash particles: using the attributed density distribution and the end-member sphericity ($\Psi_R = 0.4$) (Bonadonna et al., 2011), a particle with diameter 100 μm travels 248 km (table 2). In this study we have used the NAME + Ganser sedimentation scheme but instead considered the variation in plume height and the meteorlogical conditions during the event to calculate the maximum travel distance of Eyjafjallajökull 2010 ash particles. We show that during the first 24 hours of the eruption particles with diameter 100 μm can travel up to 3$\times$ further (figure 6) than calculated using the same sedimentation scheme with a constant plume height and wind speed (table 2). To effectively model the transport of volcanic ash clouds requires that we account for the weather that occured during the eruption.

Ash samples from the Eyjafjallajökull 2010 volcanic ash cloud were collected across Europe; in Norway, the Faroe Islands and the UK (Stevenson et al., 2012). Samples were found to have a mean diameter of 25 μm in the UK (distance from vent $\sim$1000 – 1800 km), 48 μm in Bergen (Norway) (distance from vent 1342 km) and 40 μm in the Faroe Islands (distance from vent 660 km) (Stevenson et al., 2012). The maximum particle diameter observed was 100 μm in the Faroe Islands and UK (Lincolnshire -0.38E, 52.74N, distance from vent 1646 km), and 90 μm in Bergen (Stevenson et al., 2012). Considering the first 24 hours of the Eyjafjallajökull 2010 eruption, during which time plume heights varied between 5.5 – 9 km asl, NAME simulations with the London VAAC input particle characterstics, can account for the mean measured particle size of deposits sampled across the UK and the Faroe Islands: a 30 μm particle travelled up to 1886 km and particles with diameter 100 μm reach the Faroe Islands (table 2). The maximum travel distance of a particle throughout the whole eruption can be determined from the modelled accumulated deposit, where only deposits greater than or equal to the mass of one particle in 1 m^2, which depends on the size and density of the particle, are considered. Figure 10 shows the maximum travel distance of ash particles with with the end-member sphericity (Ψ_R = 0.4) and the density distribution attributed to the Eyjafjallajökull 2010 ash particles, represented as great circle distances. A particle with diameter 30 μm and 100 μm travels up to 4711 km and 1349 km respectively (figure 10). We can now account for particles with diameter 90 μm observed in the deposits sampled in Bergen (Norway), but we are unable to account for particles with diameters of 100 μm observed in the deposits sampled in Lincolnshire (UK). Given that we have accounted for the meteorological conditions that occured during

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the eruption, including dominant periods of flow towards Europe and the UK (Petersen et al., 2012) this would suggest that we are not correctly representing the transport of these particles. Here we have assumed that the theoretical density distributions and sedimentation scheme chosen represent well the fall velocity of non-spherical, vesicular, ash particles. There are several schemes given in the literature for calculating the fall velocity of non-spherical ash particles (e.g. Wilson and Huang, 1979; Dellino et al., 2005) and a range of parameters are used to identify the shape of volcanic ash (e.g. Wilson and Huang, 1979; Riley et al., 2003; Dellino et al., 2005; Bagheri et al., Subm), further, the methods for measuring the shape parameters vary. Liu et al. (2014) analysed samples of resuspended Eyjafjallajökull and Grímsvötn ash particles deposited in Reykjavík on the 7th March 2013, they determine the sphericity of the particles from BSE-SEM images of samples mounted in polished grain mounts, Riley et al. (2003) and Bonadonna et al. (2011) use 2-dimensional projections of 3-dimensional grains. Liu et al. (2014) found Ψ_R to range from 0.56 to 0.2 for particles with diameter 20 to 160 μm, with modes at 0.25 and 0.55 for Grímsvötn and Eyjafjallajökull particles respectively. In this study we chose to use the Ganser scheme and shape parameter sphericity (Ψ_R) as defined by Riley et al. (2003) which was shown to be the best modelling practice for volcanic ash by Alfano et al. (2011). However, our results agree with the work of Stevenson et al. (2015), and suggest that the best modelling practice is under-estimating the fall velocity of non-spherical, vesicular, ash particles. A greater dependence on particle shape and density when calculating the terminal fall velocity of ash particles could increase the forecast sensitivity to the parameter values chosen.

NAME simulated PSDs of the Eyjafjallajökull 2010 ash cloud and deposits, using the particle characteristics assigned for operational runs by the London VAAC compare well to the measured PSD of the ash cloud by the FAAM research aircraft and the measured PSD of deposits on the Faroe Islands (table 3). However, both the simulated PSD of the ash cloud and deposit under-estimate the mass fraction of particles with diameter $<$ 10 μm, and over-estimate the mass fraction of particles in the size range 10 $\leq D \leq$ 30 μm, additionally the mass fraction of particles with diameter 30 – 100 μm is under-estimated when compared to the Faroe Islands deposit. This result agrees with the study of Dacre et al. (2013) who also found that using the London VAAC operational PSD in NAME under-estimated the mass of ash particles with diameter 1 – 10 μm and over-estimated the mass of ash particles with diameter 10 – 30 μm compared to the measured PSD of the ash cloud by flights conducted by the FAAM research aircraft on the 5th, 14th and 17th May 2010.

The measured TGSD of Eyjafjallajökull 2010 ash deposits from the 4 – 8th May 2010 period represents plume heights between 5.5 – 10 km asl and the dry magamtic phase of the eruption, as opposed to the earlier phreato-magmatic activity (Bonadonna et al., 2011). Using this PSD as the input, source, PSD in NAME did not improve the model fit to either the measured in-cloud PSD by the FAAM aircraft on the 17th May 2010 or the measured PSD of the Faroe Islands deposit which

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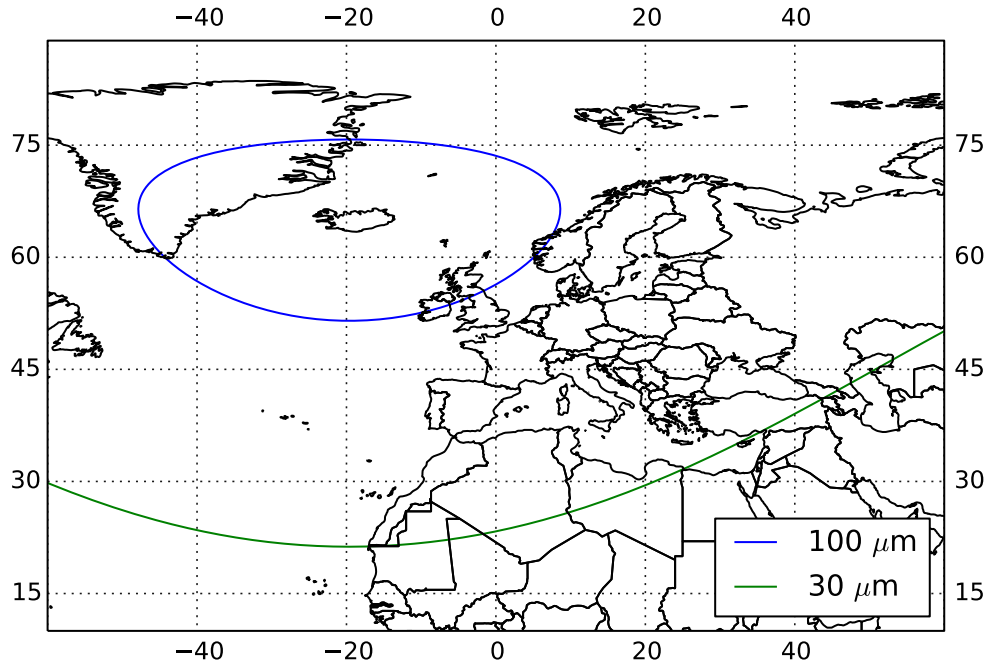

Figure 10: Maximum travel distance of an ash particle from the eruption of Eyjafjallajökull in 2010, determined from the NAME simulated accumulated deposit over the entire eruption. The particle characteristics used are the minimum measured sphericity (Ψ_R = 0.4) and density distribution attributed to the Eyjafjallajökull 2010 ash particles.

accumulated between the 14th April – 8th May 2010. When compared to the in-cloud PSD the mass fraction of particles with diameter < 4 μm was under-estimated and the mass fraction of particles with diameter 4 – 16 μm over-estimated. When compared to the Faroe Islands deposit the mass fraction of particles with diameter < 32 μm is under-estimated and the mass fraction of particles with diameter 32 – 62.5 μm is over-estimated. The deposit sampled on the Faroe Islands represents ash that accumulated between the 14th April – 8th May 2010, and includes ash from the phreato-magmatic phase of the eruption which is not captured by the deposit and MSG-SEVIRI PSD. The ash cloud was sampled by the FAAM on the 17th May 2010, at a later period than the deposit and MSG-SEVIRI PSD, however it was during the dry magmatic phase of the eruption (Gudmundsson et al., 2012) and the plume height during the preceeding 48 hours was $\sim$8 km, thus the deposit and MSG-SEVIRI PSD can be considered a good representation of the eruptive activity. Using a PSD determined from deposits close to source as the input, source, PSD to initialize a dispersion model to forecast the long range transport of volcanic ash clouds, where particles in the size range 0.1 – 100 μm are considered, should therefore be treated with care as it may not correctly represent the mass fraction of the smallest ash particles, even when combined with satellite observations.

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The PSD used operationally by the London VAAC is based upon airborne measurements of ash clouds from the eruptions of Mount St. Helens (1980), Mount Redoubt (1990) and St. Augustine (1976), given in Hobbs et al. (1991). Hobbs et al. (1991) were only able to detect particles with diameter < 48 μm, and the PSD used by the VAAC is extrapolated to 100 μm, therefore there is large uncertainty associated with this size fraction. Using the VAAC default PSD as the source, input, PSD in NAME under-estimates the mass fraction of particles with diameter 30 – 100 μm when compared to the Faroes Island deposit (figure 8) which includes particles with diameter up to 176.8 μm, and is a poorer fit to the deposit than the ash cloud sampled by the FAAM which includes particles up to 35 μm only.

Rust and Cashman (2011) have shown that the PSD of erupted pyroclasts can be grouped depending on the type of eruption. The eruption of Eyjafjallajökull in 2010 has been assigned a VEI (Volcanic Explosivity Index) of 3, which is associated with Vulcanian type eruptions (Gudmundsson et al., 2012); the eruption had a smaller magma discharge rate and plume height than a Plinian eruption, but was unusual in that it was of very long duration and resulted in ash dispersal over a much greater area than has been previously reported for an event of this magnitude (Gudmundsson et al., 2012). In figure 11 we compare the model output to the measured PSD over the southern North Sea by the FAAM using the measured PSD of the Plinian-sub Plinian Mount St. Helens 1980 eruption (Hobbs et al., 1991) and the Vulcanian type eruption of Ruapehu in 1996 (Bonadonna and Houghton, 2005). Bonadonna and Houghton (2005) determined the TGSD of the Ruapehu 1996 deposit using the Voronoi tessellation technique. Using the Ruapehu PSD as the input source PSD in NAME results in a better fit to the FAAM data (RMSE 0.0506) than the measured in-cloud PSD from the Mount St. Helens eruption (RMSE 0.0940) (table 3).

NAME does not simulate aggregation processes which preferentially remove ash particles with diameter < 125 μm from the ash cloud (e.g. Brown et al., 2012). Evidence of aggregation during the eruption of Eyjafjallajökull in 2010 includes the abundance of very fine ash beds and ash aggregates on land in Iceland (Gudmundsson et al., 2012) and observations of the fall out and deposition of ash from the ash cloud in Iceland using high speed imaging (Taddeucci et al., 2011). Correctly capturing these processes would improve the NAME simulated PSD of the ash cloud and deposits downwind. An aggregation scheme has been introduced to the Eulerian model FALL3D for the transport and deposition of volcanic ash (Costa et al., 2010; Folch et al., 2010), however Folch et al. (2010) found that the introduction of the aggregation scheme increased computational time by a factor of 2 – 4 which would make this difficult to implement operationally. An alternative approach is to represent more explicitely the process of aggregation in the input source PSD chosen. An interesting approach to be explored in the future is to determine whether an effective PSD could be identified through inversion techniques given observations of the ash cloud and deposits downwind.

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The input mass eruption rate used in the NAME simulations presented in this study was calculated using the Mastin et al. (2009) empirical relationship, which is based on observations of historical eruptions. These observations are biased towards a disproportionatly large number of big eruptions. Plumes from smaller eruptions can be more significantly effected by the atmospheric conditions, and the eruption of Eyjafjallajökull is a good example of a bent over plume which is strongly effected by the local wind conditions. In such a case the use of the Mastin et al. (2009) relationship can significantly under-estimate the mass eruption rate (Degruyter and Bonadonna, 2012; Devenish, 2013; Woodhouse et al., 2012). The method of calculating the mass eruption rate and any associated uncertainties does not affect the modelled particle size distributions (figures 7, 8 and 11), as it is the mass fraction of each size range which is considered. Figure 9 shows the differences in model output as the input particle characteristics are varied, and does not consider the best approach for representing the mass eruption rate, total column outputs would be different if a different scheme was used.

The sensitivity analysis of NAME forecasts of total column mass loadings suggest that the input source PSD used in NAME should be chosen with care. NAME simulations of the Eyjafjallajökull ash cloud using the London VAAC default input PSD (figure 1) compare well to measured PSDs of the deposit on the Faroe Islands and the ash cloud by the FAAM research aircraft over the southern North Sea (figures 8 and 7). The PSD used by the VAAC therefore represented a good default input source PSD to use in the absence of further information. Probability density functions of TGSDs of deposits derived for past eruptions at a given volcano (Biass et al., 2014) are also a useful reference which could be used as an initial ‘best-guess’ in dispersion models during an eruptive event.

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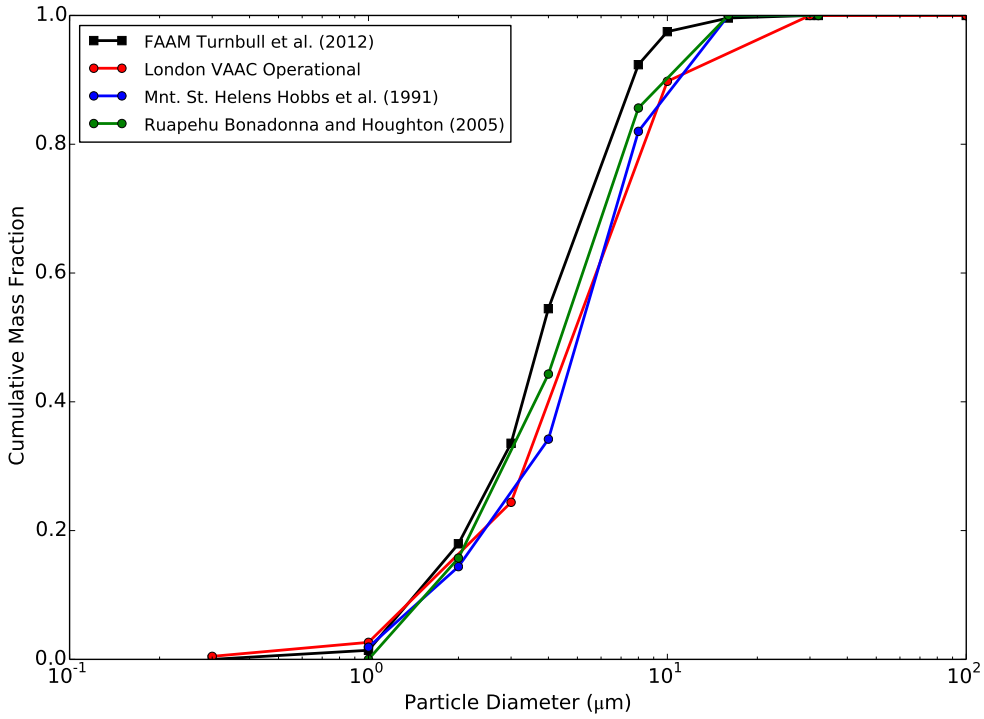

Figure 11: NAME simulated PSDs where the input source PSD is varied, compared to the PSD of the ash cloud sampled by the FAAM research aircraft, flight B530, on the 17th May 2010 during the eruption of Eyjafjallajökull 2010, 52.5 – 54.5^o N, 0 – 3^o E and between 3 – 8 km asl (Turnbull et al., 2012). Input source PSDs are: the default PSD used by the London VAAC (figure 1a), the PSD of the Mount St. Helens ash cloud measured by Hobbs et al. (1991) (figure 1b) and the measured PSD from deposits of ash from the eruption of Ruapehu in 1996 by Bonadonna and Houghton (2005).

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9 Conclusions

This study and several others which have followed the eruption of Eyjafjallajökull in 2010 (Devenish et al., 2012a,b; Webster et al., 2012), have shown that NAME is capable of capturing both quantitative and qualitative aspects of the dispersion of volcanic ash clouds. In this study we have explored the sensitivity of NAME forecasts of volcanic ash cloud transport to the input physical characteristics of the particles defined in the model. Ash particle characteristics are difficult to measure in near-real time and the London VAAC has a set of default particle characteristics that it uses in NAME in the absence of further information: ash particles are assumed to be spherical with a density of 2300 kg m^{-3} and the input, source, PSD is based on PSDs derived from airborne measurements of ash clouds by Hobbs et al. (1991). We show that the uncertainty on the calculated fall velocites, given the range of particle sphericities and density distributions typically observed for volcanic ash cloud particles are modest, $\sim$37% and $\sim$33% respectively. Using the VAAC's default particle characteristics the model simulations account for the mean measured particle diameters of ash samples collected in the UK and Faroe Islands from the eruption of Eyjafjallajökull in 2010. Accounting for the minimum measured sphericity and the particle density distribution attributed to Eyjafjallajökull ash has a negligable impact on NAME forecasts of total column ash loadings. This suggests that real-time measurements of the shape and density of ash particles would have limited benefits to forecasts of the movement of volcanic ash clouds, and a nominal value can be assigned in the operational running of NAME by the London VAAC. However, we are unable to account for the largest particle sizes observed in deposits across Europe following the eruption of Eyjafjallajökull in 2010 and Stevenson et al. (2015) have shown that using measured fall velocities rhyolitic ash particles travel up to 3$\times$ further than our best-modelling practice, this suggests that our sedimentation scheme and density distribution used may not correctly represent the fall velocity of non-spherical, vesicular, ash particles. We show that the forecasts are sensitive to the input, source, PSD defined, and the PSD used should be considered with care during operational use of the model by the VAAC. The impact of changing the PSD is similar to changing the erupted mass by over a factor of 2, and thus the uncertainty associated with the input PSD is of a similar leading order as the eruption mass uncertainty. This highlights that improved observations of the PSD of volcanic ash is critical for reducing uncertainty in quantitative forecasts of ash cloud dispersion. Determining the correct PSD to use is challenging. Using a PSD measured from deposits sampled close to the source can under-estimate the mass fraction of the smallest particles which remain in the ash cloud, and the forecast predicts lower total column ash loadings. Using a PSD measured at a significant distance downwind of the vent, even an in-cloud PSD, under-estimates the mass fraction of the largest particles which have already fallen out of the cloud, and the forecasts predict higher total column mass loadings.

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A Statistical Techniques

The Pearsons Correlation Coefficient (PCC) represents the scatter between paired (i.e. at the same location) data points, and ranges in value between -1 and +1. A value of +1/-1 represents a perfect positive or negative linear association between the two variables. In our example this accounts for the difference in both the position of the plume and the given total column mass loading values at a grid point.

The Fractional Bias represents the difference in magnitude between paired data points at the same location, it does not account for any spatial differences in the plume. Values range between -2 and +2 where positive values represent over-prediction and negative values under-predicition. A value of 0 represents a perfect match. It therefore represents the systematic bias which leads to an under/over-estimate compared to the operational model output.

The Figure of Merit in Space represents the percentage of overlap of data in space, it does not consider the magnitude of the values, in this case it represents the overlap between the position of the plume when compared to the operational model output. A high value of Figure of Merit in Space represents good model performance but a low value does not necessarily represent poor model performance, just that the plumes are shifted in space.

The Kolmogorov-Smirnov Parameter (KSP) represents the maximum absolute difference between two cumulative distributions, it is the probability of occurance of predicted and measured values not greater than a given threshold. It is a measure of how well the model reproduces the output concentrations regardless of when or where it occured. A high value suggests that the model outputs have very different output concentrations.

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